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► To cite this version:

Jean-Louis Grimaud, Dominique Chardon, Václav Metelka, Anicet Beauvais, Ousmane Bamba. Neogene cratonic erosion fluxes and landform evolution processes from regional regolith mapping (Burkina Faso, West Africa). *Geomorphology*, 2015, 241, pp.315 - 330. 10.1016/j.geomorph.2015.04.006 . ird-01419942

HAL Id: ird-01419942

<https://ird.hal.science/ird-01419942v1>

Submitted on 20 Dec 2016

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1 Neogene cratonic erosion fluxes and landform evolution
2 processes from regional regolith mapping (Burkina Faso,
3 West Africa)

4
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23 Submitted to **Geomorphology**, 7 November 2014 (MS #5234)

24 Revised manuscript, submitted 1 April 2015

31

32 **Abstract**

33 The regionally correlated and dated regolith-paleolandform sequence of Sub-Saharan
34 West Africa offers a unique opportunity to constrain continental-scale regolith dynamics
35 as the key part of the sediment routing system. In this study, a regolith mapping
36 protocol is developed and applied at the scale of Southwestern Burkina Faso. Mapping
37 combines field survey and remote sensing data to reconstruct the topography of the last
38 pediplain that formed over West Africa in the Early and Mid-Miocene (24-11 Ma). The
39 nature and preservation pattern of the pediplain are controlled by the spatial variation of
40 bedrock lithology and are partitioned among large drainage basins. Quantification of
41 pediplain dissection and drainage growth allows definition of a cratonic background
42 denudation rate of 2 m/My and a minimum characteristic timescale of 20 Ma for shield
43 resurfacing. These results may be used to simulate minimum export fluxes of drainage
44 basins of constrained size over geological timescales. Background cratonic denudation
45 results in a clastic export flux of $\sim 4 \text{ t/km}^2/\text{yr}$, which is limited by low denudation
46 efficiency of slope processes and correlatively high regolith storage capacity of tropical
47 shields. These salient characteristics of shields' surface dynamics would tend to smooth
48 the riverine export fluxes of shields through geological times.

49

50 **Keywords:** Regolith; Pediment; Landform evolution processes; Sediment routing
51 system; Source to sink.

52

53 **1. Introduction**

54 Interactions between landform evolution and regolith production and mobility
55 over shields exert first-order controls on the source, pathways and fluxes of sediments

56 and solutes over very large emerged surfaces on geological timescales (e.g., Fairbridge
57 and Finkl, 1980; Millot, 1983). Constraining these interactions on continental scales is
58 therefore relevant to quantifying the contribution of shields, as opposed to that of
59 orogens, to global sediment budgets and biogeochemical cycles in the context of long-
60 term Cenozoic climate cooling (Ollier and Pain, 1996a; Molnar, 2004; Willenbring and
61 von Blanckenburg, 2010; Goudie and Viles, 2012; Willenbring et al., 2013; Larsen et al.,
62 2014). Large-scale studies of regolith transfers are also necessary for “source to sink”
63 analyses of coupled drainage areas and sedimentary basins. The present contribution
64 aims to quantify long-term landform evolution, regolith mobility and erosional export
65 fluxes over a large region representative of shields’ sediment routing systems.

66 Tropical shields are mantled by lateritic regoliths derived from intense rock
67 weathering. Such lateritic covers are subject to remobilization by slope and fluvial
68 processes, and the reworked regoliths are commonly re-weathered after transport (Ollier
69 and Pain, 1996b). Renewed periods of regolith production by weathering and
70 remobilization by pedimentation lead to the formation of composite landscapes
71 consisting in a mosaic of lateritic paleo-landsurface remnants of various generations.
72 Such landscapes are spectacularly preserved throughout Sub-Saharan West Africa, a
73 region of more than 4.5 million km² over which a long-recognized Cenozoic regolith-
74 paleo-landsurface sequence has been dated (Beauvais et al., 2008) and correlated
75 (Beauvais and Chardon, 2013).

76 Here we develop a field- and remote sensing-based regolith-landform mapping
77 protocol applied over Southwestern Burkina Faso (Fig. 1). The investigated area is large
78 (ca. 300 x 300 km) and exposes the most widespread, type geologic and morphoclimatic
79 configuration of the West African surface (granite-greenstone terrains and flat
80 sandstones in the Guinean and Soudanian climatic zones i.e., between 10 and 13°N; Fig.

81 1). Given the constrained chronological framework of regolith-landform production
82 over the sub-region, the selected area is therefore suitable for the characterization of
83 landscape and regolith dynamics and the quantification of long-term (10^6 - 10^7 yr)
84 erosion representative of shields surfaces. Regoliths are studied here both as in-situ
85 produced or transported sediments and as paleo-landscape remnants. The obtained
86 regolith-landform map allows evaluation of the nature, distribution and preservation of
87 regolith mantles derived from a pediplain elaborated during the Early and Middle
88 Miocene (ca. 24-11 Ma) over the sub-region. Based on this map, a topographic
89 reconstruction of the pediplain is used to visualize regolith redistribution on slopes
90 during pedimentation and to evaluate landscape and drainage evolution after its
91 abandonment 11 Ma ago. Quantification of post-11 Ma dissection of the pediplain leads
92 to estimation of a type-erosion flux for shields, emphasizing the low capacity of slope
93 and alluvial processes to remove and export regolith mantles.

94

95 **2. Geomorphological and geological background**

96 *2. 1. The West African geomorphic sequence and its regoliths*

97 The following summary of the West African sequence of stepped lateritic paleo-
98 landsurfaces (Fig. 2a) is based mostly on the works of Michel (1959, 1973, 1974),
99 Eschenbrenner and Grandin (1970), Boulangé et al. (1973), Grandin (1976), Boulangé
100 and Millot (1988) (see Chardon et al., 2006; Beauvais and Chardon, 2013).

101 Each member in the sequence has a distinct regolith cover and geomorphic
102 character. The first two members of the sequence are the bauxitic and so-called
103 Intermediate surfaces, which bear thick *in-situ* formed regoliths capped by bauxites and
104 ferricretes, respectively. Bauxites are the end-product of a period of enhanced chemical
105 weathering that started in the Late Cretaceous and culminated in the Mid-Eocene.

106 Bauxites seal a topography called the African Surface, which makes the present-day
107 envelope of the West African relief. The Intermediate surface corresponds to a
108 differentiated landscape carved in the African bauxitic surface. The following three
109 stepped paleo-landsurfaces of the sequence are glacis (French term for pediments)
110 called the High, Middle and Low glacis. Glacis surfaces are commonly covered by a
111 detrital layer issued from degradation of earlier landforms. Each glacis has undergone
112 weathering after pedimentation, indicating repeated transitions from arid or semi-arid
113 pedimentation to seasonally contrasted or wet tropical weathering. Weathering periods
114 generally ended with the formation of a ferricrete cementing the glacis surfaces and
115 their detrital cover. Today, glacis occupy an overwhelming part of the West African
116 landsurface (Beauvais and Chardon, 2013; Grimaud, 2014). The sequence is best
117 preserved in the Sahelian and Soudanian climatic zones. The glacis, having undergone
118 relief inversion, show evidence of degradation further south under the humid climate of
119 the forest zone (Grandin, 1976). Though originally defined in the French-speaking
120 countries, the sequence or elements of the sequence have been formally or implicitly
121 identified and mapped in other countries of the sub-region (e.g., Fölster, 1969; Grandin
122 and Hayward, 1975; Bowden, 1987; Durotoye, 1989; Thomas, 1980, 1994; Teeuw,
123 2002; see Beauvais and Chardon, 2013; Grimaud, 2014).

124 The Lower to Mid-Eocene age of peak weathering and abandonment of the
125 bauxitic surface had long been stratigraphically bracketed (e.g., Millot, 1970). In
126 absence of any stratigraphic constraints or radiometric data, the three glacis were
127 thought to reflect Quaternary glacial – interglacial climatic cycles having led to the
128 dissection of the Intermediate landscape of supposedly Latest Pliocene age (Michel,
129 1959). Absolute age constraints on the paleo-landsurface sequence were provided by
130 radiometric ^{39}Ar - ^{40}Ar dating of supergene cryptomelane ($\text{K}_{1-2}\text{Mn}_8\text{O}_{16}, \text{nH}_2\text{O}$) around

131 Tambao in Northeastern Burkina Faso (Hénocque et al., 1998; Colin et al., 2005;
132 Beauvais et al., 2008; summarized in Beauvais and Chardon, 2013; Fig. 1). Data
133 retrieved from the weathering profiles of each surface relict define five Ar-Ar age
134 groups (ca. 59 - 45, 29 - 24, 18 -11, 7-6, and 3 Ma; Fig. 2b) bracketing periods of
135 chemical weathering of the paleosurfaces. The lower limit of these age groups
136 corresponds to the abandonment age of each paleosurface, i.e., the stabilization of the
137 weathering front by the end of the weathering period preceding duricrusting and
138 subsequent incision of each paleosurface (Fig. 2b). Abandonment ages are ca. 45 Ma for
139 the African bauxitic surface and 24 Ma for the Intermediate surface. The High glacia
140 surface developed between ca. 24 and 11 Ma with a predominance of weathering
141 between 18 and 11 Ma. In Syama (Southern Mali; Fig. 1), Ar-Ar ages on supergene
142 alunite and jarosite (Vasconcelos et al., 1994) confirm predominance of weathering
143 conditions during elaboration of the Intermediate surface until the end of the Oligocene
144 and weathering of the High glacia during the Mid-Miocene (Fig. 2b). Weathering and
145 later abandonment of the Middle and Low glacia have occurred around 7-6 and 3 Ma,
146 respectively (Fig. 2).

147

148 2. 2. Geological context

149 The study area belongs to the Paleoproterozoic portion of the West African
150 craton exposed South of the Sahara (e.g., Feybesse et al., 2006; Fig. 1). This area
151 comprises Birimian (2.2-2.1 Ga) granite-greenstone terrains over two-thirds of its
152 surface and the Neoproterozoic sandstones of the Taoudeni basin, which were intruded
153 by dolerite sills (Fig. 3). The basement comprises the Banfora, Houndé and Boromo
154 greenstone belts and intervening granitoids (Baratoux et al., 2011; Metelka et al., 2011;
155 Fig. 3b). Greenstones consist mainly of mafic volcanics (basalts and andesites) and

156 volcanosediments, whereas granitoids comprise TTG (tonalite - trondhjemite –
157 granodiorite) and granitic plutons. The escarpment delimiting the Banfora plateau marks
158 the southeastern boundary of the Taoudeni basin (Fig. 3a). The southern part of the
159 plateau bears the highest summits of the region and the height of the escarpment
160 decreases towards the NE. The study area mostly belongs to the Mouhoun drainage
161 system (the main stream of the Volta drainage system, i.e., the Black Volta) and is
162 fringed by the catchments of the Upper Niger River to the west, the Comoé River to the
163 southwest and the Nazinon River to the east (part of the Nakambé drainage system i.e.,
164 the White Volta; Figs. 1, 3).

165

166 2. 3. Earlier works on the study area

167 The pioneering study of Daveau et al. (1962) around Houndé and Gaoua (Fig. 3)
168 led to the identification of stepped lateritic landsurfaces and to the description of what
169 will be recognized later as the High glaciais. Later on, the southern part of the study area
170 and its extension in the Ivory Coast became one of the first regions where the entire
171 West African regolith-landform sequence was thoroughly documented (Eschenbrenner
172 and Grandin, 1970). Boeglin and Mazaltarim (1989) and Boeglin (1990) later focused
173 on an area of ca. 50 x 30 km around Gaoua (Fig. 3), where they mapped the High glaciais
174 relicts and performed a detailed geochemical study of their ferricretes. More recently,
175 Bamba (1996) and Bamba et al. (2002) have refined detailed mapping of the three glaciais
176 around the Poura mine (left bank of the Mouhoun River, West of Tô; Fig. 3) to study
177 surface remobilization of gold. Lately, Butt and Bristow (2013), reporting on gold
178 prospects of the area, argued for the stepped character of the glaciais and the detrital
179 nature of their ferricretes, which had been documented by geomorphologists since the
180 late 1950's (e.g., Michel, 1959; Daveau et al., 1962) but largely ignored since then.

181

182 **3. Field relationships and regional-scale regolith-landform mapping**

183 3. 1. Rationale

184 As opposed to the other regolith-landform associations of the lateritic paleo-
185 landsurfaces sequence, High glacia remnants have both distinctive field and remotely-
186 sensed mappable properties at the resolution of the digital data and given the size of the
187 map area (see below). Our work therefore focused on the High glacia regolith-landform
188 associations, which were characterized and mapped for evaluation of the pedimentation
189 and weathering processes controlling the development of the High glacia paleo-
190 landsurface. The topography of this paleo-landsurface was then reconstructed as a
191 datum for the quantification of regional erosion after its abandonment. Higher paleo-
192 landsurface relicts of the sequence (i.e., bauxitic and Intermediate) were already
193 inherited landforms incorporated to the High glacia landscape and are still preserved in
194 the current landscape. As seen below, their duricrusts are also found as reworked
195 elements in the High glacia regoliths. Field characteristics of the higher landsurfaces are
196 therefore described and their occurrences reported in the regolith-landform map. Relicts
197 of the Middle and Low glacia (i.e., lower landsurfaces) occupy lower parts of today's
198 landscape that have been excavated in the High glacia surface. Their field characteristics
199 are briefly described below.

200

201 3. 2. Field typology of landforms and associated regoliths

202 3. 2. 1. Field survey

203 Our field observations were carried out at stations (Fig. 3b), which consist of
204 areas of several km², where landscape analysis was undertaken by interpreting sceneries
205 using 1/50 000 and 1/200 000 scale topographic maps. Mapping of regolith-landform

206 units was undertaken at key stations. Landscape interpretations/chronologies and
207 identification of paleo-landsurface relicts were complemented by in-situ examination of
208 the regolith along tracks through the landscape and at road cuts. This work included the
209 identification of the textures of the exposed lateritic weathering profile horizons and the
210 description of the sedimentary facies of the transported regoliths.

211

212 3. 2. 2. Higher landsurfaces

213 Bauxite relicts form mesas on the highest summits. On the basement, those
214 mesas are restricted to mafic substrates and their size is generally limited (i.e., less than
215 100 m wide) (Fig. 4a). In the Taoudeni basin, bauxitic plateaus are larger and mostly
216 restricted to dolerites (Figs. 4b, 4c). The bauxites have typical pink-whitish pisolithic
217 textures (Fig. 4d) or less evolved, nodular-mottled textures (Fig. 4e). In the Taoudeni
218 basin, particularly on dolerites, the Intermediate ferricrete caps gently dipping convex-
219 concave surfaces forming the downslope extension of the bauxitic plateaus (Figs. 4b,
220 4c). These ferricretes (Figs. 4f, 4g) show gradational variation in composition and
221 texture with the bauxites. This is interpreted to result from iron leaching that leads to
222 relative aluminum enrichment of the bauxites and net concomitant iron accumulation in
223 the ferricretes (Boulangé, 1986). Intermediate ferricretes on sandstones display typical
224 nodular textures (Fig. 4g). On the basement, the Intermediate ferricrete is rarely found
225 in-situ but rather as pebbles or cobbles reworked in the High glacia ferricrete (see
226 below). Instead, a residual “Intermediate relief” surrounds bauxitic duricrusts and may
227 be locally paved with a cemented scree of bauxite cobbles (Fig. 4a). Bauxites are
228 underlain by weathering profiles that are at least 80 m thick, as estimated on the
229 stripped slopes of mesas, whereas Intermediate weathering profiles exceed 20 m in
230 thickness.

231

232 3. 2. 3. High glacis

233 The best visible High glacis relicts are kilometer-scale plateaus (the largest ones
234 being concave) shallowly dipping away from the topographic massifs made of mafic
235 volcanics, which may carry higher relict paleo-landsurfaces (Fig. 5a). Such High glacis
236 relicts are often separated from the massifs by a peripheral hollow due to incision of
237 their upslope parts (see also Beauvais et al., 1999). This configuration indicates that the
238 High glacis has occupied large piedmonts that have been dissected. Dissection of the
239 High glacis ferricrete has carved 1 to 20 m-high scarps dominating slopes cut into soft
240 regoliths and covered by ferricrete debris (Figs. 5, 6a, 6b).

241 Most High glacis plateau ferricretes consist of a conglomerate ranging from
242 gravel to cobble, with a predominance of gravels mostly made of ferruginous nodules
243 (Fig. 6). Matrix silts and sands may bear quartz, particularly on volcano-sediments and
244 granitoids. Coarse gravels and cobbles are made of Intermediate-type ferricretes, vein
245 quartz and bauxites (Figs. 6a-6e). Conglomerates are commonly matrix- or block-
246 supported, suggesting emplacement by mudflow or debris flow process. Channels are
247 observed at the base of the ferricrete and decimeter- to meter-scale oblique
248 stratifications are recognized in the downslope parts of some large High glacis relicts
249 (Fig. 6d). The coexistence of mud or debris flow and alluvial facies in the High glacis
250 cover is indicative of interplay of slope and fluvial processes during pedimentation
251 (Dohrenwend and Parsons, 2009).

252 The noticeable absence of a mottled horizon under the glacis conglomerate or
253 exposure of deep portions of weathering profiles right under that cover (Figs. 6a, 6c)
254 indicates that these weathering profiles have been truncated before deposition of the
255 conglomerate. Duricrusting (i.e., iron impregnation or cementation) is mostly restricted

256 to the conglomerates cover (Figs. 6a, 6b). Still, soft ferricretes, i.e., pedogenic
257 impregnation of a mottled clay lateritic horizon (Tardy, 1997), are common under the
258 cemented conglomerates. This suggests renewed weathering along the upper fringe of
259 the truncated lateritic profiles under the conglomerates. At certain locations, the
260 ferricrete made of cemented debris flows may even be underlain by a type-succession of
261 weathering profile horizons (i.e., soft ferricrete, mottled layer, fine saprolite, coarse
262 saprolite; Figs. 6e, 6f), suggesting the full development of a weathering profile after
263 pedimentation.

264 On granitoids located away from topographic massifs (e.g., west of Djigoué and
265 south of Bobo Dioulasso, Tô region), the High glacia ferricretes are generally thinner
266 (0.5 - 2 m) and mantle smooth interfluves (Figs. 5c, 5d). Such ferricretes have
267 vermicular and nodular textures and are devoid of clasts. They appear less ferruginous
268 and less indurated than their plateau counterparts. These ferricretes show gradational
269 contacts with underlying soft ferricretes and are part of in-situ weathering profiles.

270

271 3. 2. 4. Lower landsurfaces

272 The restricted size of higher paleo-landsurfaces remnants is primarily due to the
273 development of the High glacia. But rejuvenation of the slopes of the residual reliefs
274 carrying those remnants and the formation of peripheral hollows (Fig. 5) is attributed to
275 the following periods of glacia formation. The landsurface between High glacia relicts
276 and the modern alluviums is almost exclusively occupied by the Middle and Low glacia.
277 These glacia may be stepped or may combine to form a single polygenic landsurface.
278 The regolith of the Middle and Low glacia is comparable to that of the High glacia, but
279 their ferricrete is generally thinner to absent and may be covered over large areas by
280 loose material made of reworked saprolite, sand and ferruginous gravels.

281

282 3. 3. Regional-scale regolith-landform mapping protocol

283 On basement terrains, High glacis units were manually mapped in GIS on the
284 basis of their airborne gamma ray spectrometry and their photo-interpretation signature
285 (Google Earth), both validated by the field survey / maps (Figs. 3, 7). Airborne gamma
286 ray spectrometry was acquired during the 1998–1999 SYSMIN (System for Mineral
287 Products) project (Metelka et al., 2011). Those data display uranium (U), potassium (K)
288 and thorium (Th) contents of the upper 30 cm of the Earth surface (Dickson and Scott,
289 1997). The spatial resolution of the data on the basement is 125 m (with an original line
290 spacing of 500 m). Spatial resolution of 250 m (line spacing of 1000 m) over the
291 Taoudeni basin has required resampling to 125 m in the area to obtain a harmonized
292 image. Both the raw and enhanced (Th/K ratio) gamma-ray spectrometry data were
293 used (Fig. 7) during the analysis and combined with the shaded relief maps of the
294 SRTM (Shuttle Radar Topographic Mission) data (90 m spatial resolution). The High
295 glacis ferricretes have a distinctive photo-interpretation texture that spatially matches
296 blue-green colors in raw gamma-ray data. This attests of a strong depletion in potassium,
297 and a relative enrichment in thorium and uranium as a result of weathering processes
298 (Dickson and Scott, 1997; Wilford et al., 1997; Metelka et al., 2011). However, uranium
299 has a complex behavior during weathering and regolith formation (Dickson and Scott,
300 1997; Wilford et al., 1997; Dequincey et al., 2002). Therefore, the Th/K ratio was
301 preferred to further enhance the signature of the High glacis ferricretes (in yellow-red)
302 and remove any regional gradients in the data (Figs. 7b, 7d).

303 The spatially homogeneous radiometric signature of the High glacis relicts
304 suggests that their ferricretes have been geochemically homogenized by weathering
305 despite the variability of their substrate. Indeed, trace element concentrations

306 systematically converge as a function of the Fe_2O_3 content of the ferricrete taken as an
307 index of their geochemical maturity (Boeglin and Mazaltarim, 1989; Boeglin, 1990).
308 Conversely, the ambiguous radiometric signature of the Middle and Low glacis may be
309 interpreted as a lower maturity of their ferricrete and a mixed signature of the material
310 covering their ferricrete.

311 The strong Th and exceptionally low K signatures of the Taoudeni basin (Figs.
312 7a, 7b) are explained by the mineralogy of the sandstones that are originally quartz rich
313 and correlatively poor in clays. The larger pixel size (250 m) of the gamma ray
314 spectrometry may have further modified the Th signature of High glacis ferricretes by
315 sampling other surface material with high Th content (e.g., Bauxite and Intermediate
316 duricrusts). Therefore, the high values (red colors) of Th/K ratio image (Fig. 7b) may
317 not only be due to High glacis ferricretes. Hence, regolith-landform mapping over the
318 Taoudeni basin was mainly based on photo-interpretation in Google Earth and field
319 observations.

320 Bauxite and Intermediate surface relict occurrences were mapped by photo-
321 interpretation, field observations and after Petit (1994). River plain sediments were
322 mapped mainly on the basis of their negligible relief and the characteristic geometries of
323 stream networks in the gamma-ray imagery. Because of the higher mobility of U
324 compared to Th, river plain sediments raw gamma-ray signature may indeed vary i.e.,
325 color of the source material (variable content of radioelements), very light to white color
326 (high U / Th / K signal), or very dark color (low U / Th / K signal). It might reflect the
327 composition of the source material, sediment mixing, or strong attenuation by
328 vegetation or water (Fig. 7a).

329

330 3. 4. Interpretation of the High glacis regolith-landform map

331 For the purpose of the interpretation, the mapped regolith-landform units were
332 superimposed on the topography in Fig. 8 and on the geological map from Metelka et al.
333 (2011) in Fig. 9. Comparison of these two maps highlights major contrasts in the pattern
334 of High glaciae relicts' distribution on either side of the main watersheds, even on the
335 same geological substrate. This is particularly exemplified on TTG across the
336 southwestern and eastern Mouhoun watersheds (Figs. 8, 9a). Density of High glaciae
337 relicts generally decreases when approaching the main streams (Fig. 8). Relicts are best
338 preserved on TTG and granite (45 and 20% of the total surface of the relicts,
339 respectively). Greenstones underlay less than 25% of the total area of the preserved
340 High glaciae ferricrete (Figs. 3b, 9c). Fifteen percent of the map area of the TTG and
341 more than 20% of the granitoids are covered by High glaciae ferricretes, whereas only
342 10% of the greenstones are covered (Fig. 9d).

343 Type-patterns of High glaciae regolith-landforms may be distinguished in the
344 study area (Figs. 5, 9, 10). In high relief volcanic terrains, High glaciae relicts are well
345 preserved on piedmonts (Fig. 10a), equally on greenstones and granitoids. Volcano-
346 sedimentary-granitic terrains of moderate relief (e.g., Djigoué area) display High glaciae
347 relicts that once coated wide N-S trending valleys, which were in turn dissected by
348 second-order narrower E-W oriented valleys (Fig. 10b). Over TTG terrains of the
349 Comoé catchment (e.g., Dandougou-Bobodioulasso area), a very low relief and very low
350 slope High glaciae surface is preserved on sparse residual hills where the ferricrete may
351 be dismembered into boulders resting on the soft ferricrete (Fig. 10c). The resulting
352 landscape is a convexo-concave plain of low amplitude that is gently dipping towards
353 the main stream (Fig. 10c). Equivalent TTG terrains in the Nakambe catchment reveal a
354 better-preserved High glaciae paleo-landsurface (see above), with planar to weakly
355 concave relicts delineating a plain of very low-relief surface envelope (Fig. 10d).

356 The occurrence of strong and thick plateau ferricretes upon granitoids on the
357 piedmonts of volcanic reliefs contrasts with the weak and thin ferricretes preserved on
358 the same granitoids of the low land (Figs. 5a-5d). This reflects the high iron content and
359 thickness of the piedmont conglomerates made of reworked bauxitic and Intermediate
360 duricrusts that formed on iron-rich mafic rocks. On the contrary, low land ferricretes are
361 solely issued from in-situ weathering of iron poor granitoids. The limited lateritic cover
362 of mafic volcanics massifs further argues for the transfer of upland iron-rich duricrusts
363 as colluviums to the High glacis piedmonts. This transfer has resulted in net iron
364 accumulation on glacis surfaces further down in the landscape (Michel, 1973, 1974;
365 Beauvais et al., 1999).

366 The above analysis indicates that the High glacis landsurface was a pediplain,
367 defined as a surface of coalescent pediments (Maxson and Anderson, 1935). The spatial
368 distribution of these pediments and their regolith as well as the preservation pattern of
369 the High glacis pediplain are spatially controlled at various scales. A first, local control
370 of this distribution is exerted by the lithology of the bedrock. This sets the relief and
371 therefore the pedimentation process, as well as the weathering mode of the pediplain.
372 On a regional scale, pediplain dissection and erosion mode appear to have been
373 compartmentalized among drainage basins.

374

375 **4. Reconstruction of the Late Mid-Miocene pediplain and quantification of its** 376 **erosion**

377 4. 1. Methodology

378 The high density of High glacis relicts warrants reconstruction of the pediplain
379 they were part of. The reconstructed topography is a triangulated surface interpolated
380 from a set of geo-referenced points generated from the polygons of the High glacis

381 relicts using the DSI method (Gocad software; Mallet, 1992). Each High glacia polygon
382 was converted to a grid of 500 m-spaced points inside its limits. The elevation of the
383 SRTM digital elevation model, degraded to 500 m resolution, was then sampled for
384 each point to form the High glacia relicts' elevation dataset. The interpolated surface
385 was created from this dataset through a series of iterations. The number of surface's
386 triangles, which reflected the distribution of the dataset for the first incremental surface,
387 was increased at each iteration step, while the High glacia surface was forced to remain
388 above the modern topography. This allowed for the inclusion of the reliefs dominating
389 the pediplain and preserved the composite nature of the High glacia landscape. The
390 resulting topography is shown on Fig. 11. Erosion post-dating the abandonment of the
391 pediplain was obtained by subtracting the modern topography from that of the
392 reconstructed pediplain (Figs. 11a-c). The drainage network and a slope map were then
393 automatically extracted from both the modern and pediplain digital elevation models for
394 comparison (Figs. 11d-h).

395

396 4. 2. Late Mid-Miocene topography and landscape dynamics

397 The pediplain displays the same first-order relief pattern as the current
398 topography. The main divides such as the inselbergs and the main escarpment already
399 existed at the time the pediplain was functional and did not migrate since then (Figs.
400 11a-b). The main river network was therefore already established since the Late Mid-
401 Miocene with the exception of a few local river rearrangements (Figs. 11d, 11h). Our
402 reconstruction does not allow for testing whether the Upper Mouhoun (flowing on the
403 Taoudeni basin) was connected to the lower Mouhoun at the time of the High glacia
404 (Fig. 11h). This rearrangement (Hubert, 1912; Palausi, 1959) occurred after the upper
405 Mouhoun River, which used to flow northeastward into the Niger River since the

406 Eocene, formed the Gondo Plain internal delta (Beauvais and Chardon, 2013; Grimaud
407 et al., 2014; Fig. 1).

408 The spatial variability of the pediplain relief on the basement is representative of
409 the type-landscapes shown in Fig. 10 (comparisons with Figs. 9a, 12). One distinguishes
410 (i) an inselberg-studded pediplain around volcanic terrains, (ii) almost flat lands over
411 TTG, and (iii) a mix of the first two landscape types over volcano-sediments and
412 associated granite plutons. In the Taoudeni basin, the pediplain consists of wide and
413 shallowly dipping piedmonts connecting sandstone or higher landsurfaces plateaus to
414 the main streams.

415

416 4. 3. Interpretation of the pediplain dissection pattern

417 The pediplain has a low drainage density compared to that of the modern
418 landscape, with typical pediment widths of 10-20 km (up to 30 km over the sandstones;
419 Figs. 11, 12). Dissection of the High glaciais landscape accompanied an increase in the
420 drainage density. The modern valley sides rarely exceed 5 km in width (Figs. 3, 8).
421 Dissection of the pediplain also resulted in an increase in slopes (Fig. 11g), which is in
422 agreement with the evolution from a pediplain towards narrower / deeper valleys. In
423 other words, the Middle and Low glaciais, which contributed to the dissection of the
424 pediplain, never attained the width of the High glaciais.

425 Post-11 Ma dissection of the pediplain is non-uniformly distributed, with
426 denudation depths ranging from 0 to 100 m and corresponding denudation from 0 to 10
427 m/my (Fig. 11c). Extreme denudations (> 60 m, > 5 m/My) over basement terrains are
428 concentrated on the steep slopes of residual reliefs (Figs. 6, 10, 11). Along the two main
429 streams, denudation is moderate to high (30 - 60 m; 2.5 - 5 m/My). The denudation is
430 rather uniformly distributed in the Comoé drainage, meanders being delineated by

431 denudation patterns in the vicinity of the Mouhoun River (Fig. 11c). The lowest erosion
432 range (< 10 m, < 1 m/My) is recorded east of Tô, southeast of Bobo-Dioulasso and
433 around Gaoua and Djigoué (Fig. 11c), in agreement with the high degree of preservation
434 of the pediplain relicts (Fig. 8). The volume eroded from the pediplain over the map
435 area represents a spatially averaged denudation of ca. 20 m, corresponding to a
436 denudation rate of about 2 m/My.

437 Dissection of the Late Miocene pediplain affected the whole region and did not
438 propagate upstream by scarp or knickzone retreat as advocated by King (1962) and as
439 implicitly included in river profile inversion works based on stream power incision
440 models (e.g., Paul et al., 2014). As an illustration, the bounding escarpment of the
441 Taoudeni basin is stable since at least 11 Ma (Fig. 11) and most probably 45 Ma (see
442 Grimaud et al., 2014). Higher erosion rates coincide with the main drains outside the
443 High Mouhoun, suggesting a control of the stream power (e.g., Whipple and Tucker,
444 1999) on pediplain dissection. Moderate stream power in the rest of the river network
445 would have preserved the pediplain from complete stripping. Scarps bounding High
446 glaciais relicts often coincide with second-order divides of the modern topography,
447 indicating a component of slope retreat in the dissection of the pediplain. Meandering
448 patterns of the Mouhoun River (Figs. 8, 11c) further suggest that channel migrations
449 may have contributed to slope retreat. The typical retreat distance of the High glaciais
450 relicts' edges over the study area (1-10 km) suggests a retreat rate of 100-1000 m/My.
451 Therefore, under the erosion dynamics at play since ca. 10 Ma and assuming an initial
452 valley side width of 20 km (Fig. 12), a minimum of 20 My would be required to entirely
453 resurface the High glaciais pediplain.

454

455 **5. Discussion**

456 Our results complement denudation and incision rates estimated from
457 differential elevation of sparse High glacis remnants and local base levels that have
458 mean values of 5-7 m/My but non-uniform distributions over the sub region from 2 to
459 15 m/My (Beauvais and Chardon, 2013; Grimaud et al., 2014). This suggests a
460 partitioning of erosion among geomorphic provinces, with implications on the spatial
461 variability of the sediment routing system. This variability may be a function of the
462 regional topographic relief and/or positive epeirogeny, which seem to focus higher
463 denudation (Beauvais and Chardon, 2013; Grimaud et al., 2014). Importantly, this
464 variability is also and primarily due to contrasted river network evolution among sub-
465 drainage basins separated by stationary knickzones (Grimaud et al., 2014; this work).

466 Our approach has consisted in regional scale, volumetric quantification of
467 denudation from a dated paleo-landscape datum, which enhances accuracy of
468 denudation measurement over a West African province of lowest Neogene erosion. The
469 obtained averaged denudation rate of 2 m/My is close to the lower limit of Cenozoic
470 cratonic denudation rates measured worldwide (Beauvais and Chardon, 2013). Given
471 the relevance of the size and morpho-geological context of the study area, we propose
472 that this rate determines the background denudation ‘noise’ of shields. The Late
473 Neogene increase in clastic sedimentation documented worldwide and particularly in
474 African deltas such as that of the Niger (Séranne, 1999; Jermannaud et al., 2010; Fig. 1)
475 should however be somehow included in the 2 m/My of background denudation
476 averaged since the Early Late Miocene. This issue constitutes a sizable research
477 challenge.

478 The material derived from dissection of the pediplain and removed since the
479 Late Mid-Miocene was made of regolith, i.e. the High glacis cover and its underlying
480 laterites. Indeed, no outcrop of significant surface / height has been exhumed since the

481 abandonment of the pediplain. The fact that only the main stream channels locally flow
482 atop the bedrock and that an averaged denudation of less than 20 m since abandonment
483 of the pediplain is estimated further argues for denudation being restricted to the
484 regolith. Denudation is however not strictly weathering-limited because if glacis
485 typically cut preexisting and potentially old regoliths (i.e., Late Cretaceous to Eocene
486 and Oligocene for the bauxitic and Intermediate weathering episodes, respectively,
487 Figure 2), they rarely attained the bedrock. Therefore, although weathering is
488 instrumental in producing regolith mantles available to future stripping (Fairbridge and
489 Finkl, 1980), low abrasion power of the pedimentation process combined with a
490 probably limited transport capacity of the river network should be regarded as the
491 limiting factor of tropical shields denudation.

492 The minimum characteristic timescale of 20 My for shield resurfacing indicated
493 by post-Mid-Miocene drainage growth reinforces the view that shields are non-
494 equilibrium landscapes, which increase their relief through geological time by
495 preserving old landforms (Thomas, 1989; Twidale, 1991; Bishop, 2007; Beauvais and
496 Chardon, 2013). A long resurfacing timescale also implies the high sediment retention
497 capacity of shields as those sediments stem from weathering mantles dating back from
498 at least the Early Paleogene and have been slowly recycled through the landscape since
499 then. This suggests that weathering and slope processes play a major role, comparable
500 to or greater than sediment transport by rivers (Métivier and Gaudemer, 1999;
501 Jerolmack and Paola, 2010) in “shredding” and particularly smoothing or buffering
502 climatic signals in the sedimentary record. Still, the ubiquitous character of the West
503 African lateritic paleo-landsurfaces sequence and its chronology would suggest that
504 Cenozoic landscape rejuvenation and sediment fluxes were punctuated by major
505 climatic periods. Subdued clastic exports are indeed expected for time intervals of

506 enhanced weathering during the Lower and Mid-Eocene, Late Oligocene, Mid Miocene,
507 Latest Miocene and Latest Pliocene (e.g., Fig. 2b).

508 The background denudation rate of 2 m/My may be converted into a volumetric
509 export rate of $2 \times 10^{-3} \text{ km}^3/\text{km}^2/\text{My}$ and a clastic yield of $4 \text{ t}/\text{km}^2/\text{yr}$, considering an
510 effective grain density of $2,500 \text{ kg}/\text{m}^3$ (estimated from a bulk density of $2,000 \text{ kg}/\text{m}^3$
511 and a 20 % porosity) for the stripped regolith. This value corresponds to the lower limit
512 of modern West African solid river loads ($3.5 - 300 \text{ t}/\text{km}^2/\text{yr}$) and to the $4 \text{ t}/\text{km}^2/\text{yr}$
513 measured at the outlet of the Volta catchment (Milliman and Farnsworth, 2013).
514 Therefore, cratonic erosion fluxes derived from a 2 m/My denudation rate appear to be
515 realistic for low-capacity shield drainage systems on geological timescale. Conversely,
516 this could suggest that the lower limit of large tropical shields catchments yields is
517 governed by a cratonic background denudation noise. Importantly, background cratonic
518 denudation of 2 m/My may be used to simulate the minimum export fluxes of drainage
519 basins of constrained size over geological timescales.

520

521 **9. Conclusions**

522 Regional regolith-landform mapping allows characterizing surface dynamics,
523 weathering patterns, paleotopography and dissection mechanisms of the last pediplain
524 formed over West Africa from Early to Mid-Miocene. The nature and preservation /
525 dissection patterns of the pediplain are controlled by the spatial distribution of bedrock
526 lithologies and are partitioned between large drainage basins. Quantification of the post
527 Mid-Miocene dissection of the pediplain determines a cratonic background denudation
528 noise of 2 m/My and a minimum shield resurfacing characteristic timescale of 20 My.
529 The minimum clastic yield of the major African catchments seems to be currently
530 dominated by such a cratonic noise. Our results, once combined with landscape /

531 weathering chronologies, point to the shields' high storage capacity of sediments,
532 consisting exclusively of regoliths that have been mainly recycled in the landscapes
533 since the Late Cretaceous. Such a slow regolith turnover is due to the low efficiency of
534 slope denudation processes by pedimentation and would tend to smooth riverine export
535 fluxes of shields over geological time scales.

536

537 **Acknowledgments**

538 This work was funded by WAXI, the CNRS and the ANR TopoAfrica (ANR-
539 08-BLAN-572 0247-02). The manuscript benefited from the comments of R. Teeuw
540 and an anonymous referee as well as from editorial suggestions by A. Plater. We thank
541 M. Jessell, D. Rouby, L. Baratoux and D. Huyghe for discussions and support and A.
542 Fofana for participation in the fieldwork. We acknowledge AMIRA International and
543 the industry sponsors, including AusAid and the ARC Linkage Project LP110100667,
544 for their support of the WAXI project (P934A) as well as the Geological Surveys /
545 Departments of Mines in West Africa as sponsors in kind of WAXI.

546

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713

714 **Figures Captions**

715

716 **Figure 1:** Simplified geology of sub-Saharan West Africa. The study area is shown by a
717 red frame (map modified from Feybesse et al., 2006).

718

719 **Figure 2:** (a) Lateritic paleo-landscape sequence and incision chronology of West
720 Africa (compiled after Michel, 1973; Beauvais et al., 2008; Beauvais and Chardon,
721 2013 and modified after Grimaud et al., 2014). (b) Synthesis of Ar-Ar age data obtained
722 on supergene Mn oxides. Colored time slices represent the main periods of weathering
723 correlated with the successive paleolandscapes, based on dating of cryptomelane in
724 Tambao (after Beauvais et al., 2008; Beauvais and Chardon, 2013). Ages reported from
725 Syama were obtained on alunite and jarosite (after Vasconcelos et al., 1994). Note that
726 only alunite and jarosite ages with error of less than 5 Ma are reported. Terminology of
727 the paleo-landscapes: Bx – Bauxite; Int – Intermediate; HG – High glacis; MG –
728 Middle glacis; LG – Low glacis. Syama and Tambao are located on Fig. 1.

729

730 **Figure 3:** Topography and simplified geology of the study area (see Fig. 1 for location).
731 Shaded topography (a) based on SRTM digital elevation model smoothed to 500 m
732 resolution. Simplified geology (b) is adapted from Baratoux et al. (2011). Field stations
733 are shown by yellow dots.

734

735 **Figure 4:** Illustrations of field characteristics of the higher lateritic paleo-landscapes
736 in Southwestern Burkina Faso. (a) Bauxite-capped greenstone massif, 15 km SE of
737 Iridiaka (location on Fig. 3; see also Fig. 9b). The bauxitic plateau is ca. 300 m long. (b)
738 – (c) South-looking Google Earth view and interpretation of a dolerite topographic
739 massif near Kokoro, Taoudeni basin (location on Figure 3; Vertical exaggeration x3).

740 HG - High glacia. (d) Close-packed pisolithic texture of the Iridiaka bauxite shown in
741 (a). (e) Nodular - mottled texture of the Kokoro bauxite shown in (b) and (c). (f)
742 Massive nodular facies of the Intermediate ferricrete shown in (b) - (c). (g) Iron
743 nodulation at the root of the Intermediate ferricrete over sandstones, Taoudeni basin
744 north of Banfora (location on Fig. 3).

745

746 **Figure 5:** Illustrations of the High glacia relict landforms. (a) - (b) Google Earth view
747 and interpretation of the eastern piedmont of a volcanic inselberg near Dano (Fig. 3). (c)
748 - (d) Google Earth view and interpretation of parse High glacia relict hills resulting from
749 dissection by a dendritic river network on granitic substrate, Dandougou area (Fig. 3).
750 High glacia relicts are shown in green. Dip symbols indicate High glacia
751 paleolandscape direction and sense of slope. (e) Gently west-dipping plateau bounded
752 by a 10-15 m high scarp (5 km West of Gaoua, view looking SW). (f) Closer view of
753 the plateau edge shown in (e). (g) Scarp carved in >10 m thick ferricrete cementing
754 glacia cover material (in between Djigoué and Gaoua).

755

756 **Figure 6:** Illustrations of High glacia regoliths. (a) Base of a High glacia ferricrete
757 truncating a granitic weathering profile (Djigoué area). Light-colored cobbles in the
758 ferricrete are quartz clasts. (b) Details of a High glacia ferricrete showing cobbles of
759 Intermediate ferricrete and quartz in a finer grained conglomerate, base of the cliff
760 shown in Fig. 5g. (c) Cemented debris flow on dolerites near Kokoro (site of Fig. 4b,
761 4c). Light-colored debris consists of bauxite clasts. The flow rests upon a truncated
762 weathering profile developed from dolerites that remain as core stones in the saprolite.
763 (d) Alluvial channel at the base of High glacia ferricrete carved into weathered
764 sandstones of the Taoudeni basin (South of Bobo-Dioulasso). (e) Cobble of

765 Intermediate ferricrete cemented in a High glacia ferricrete. (f) Weathering profile
766 underlying that same ferricrete (northeast of Dédougou). See Fig. 3 for locations.

767

768 **Figure 7:** Ternary image of the airborne gamma-ray spectrometry survey (a) and the
769 enhanced Th/K ratio image of the survey (b) over the study area. (c) - (f) Details of the
770 datasets that were combined for the identification and mapping of High glacia relicts on
771 the example of the Djigoué area. (c) Airborne gamma-ray spectrometry 100-250 m
772 (resampled to 100 m) resolution ternary image. The area corresponding to the High
773 glacia ferricrete has a strong thorium signature in green and a medium uranium
774 signature in blue. In this area, the substrate has a strong potassium signature in red. (d)
775 Ratio image of thorium over potassium (100 m resampled resolution). The High glacia
776 ferricrete, enriched in thorium and depleted in potassium, is highlighted in white to red
777 colors. (e) 15-m resolution LANDSAT image. (f) Mapping by construction of the
778 polygons (crossed frame) corresponding to the High glacia ferricrete, superimposed on
779 the SRTM topography. Field stations are shown as yellow dots. Rivers are shown in
780 white.

781

782 **Figure 8:** Regolith-landform map superimposed on the STRM digital elevation model
783 (90 m resolution). Black boxes correspond to the frame of the maps shown in Fig. 10.

784

785 **Figure 9:** (a) Regolith-landform map superimposed on the geology that is simplified
786 from Metelka et al. (2011). (b) Synthetic regional cross-section of the study area. The
787 line of section is located in (a). Trace of the HG surface is shown in dark green (dashed
788 where eroded). (c) Histogram of the High glacia ferricrete total area partitioned on each
789 substrate. (d) Proportion of the High glacia surface on each lithology covered by

790 ferricrete. For the whole Birimian basement, this percentage is of about 11 %. Dolerites
791 and sandstones were not distinguished in the Taoudeni basin.

792

793 **Figure 10:** Maps and cross-sections of type-landscapes on basement terrains. Frames of
794 the maps are located on Fig. 8. Dip symbols represent the direction and sense of slope
795 of the High glacis relicts (in green). Dashed lines on the maps are lithological
796 boundaries (see Fig. 9a).

797

798 **Figure 11:** (a) Modern topography based on SRTM digital elevation model smoothed to
799 500 m resolution. (b) Reconstructed topography of the High glacis pediplain. (c)
800 Denudation map obtained by subtracting present-day topography from that of the High
801 glacis stage. (d) River network automatically extracted from the digital elevation model
802 of the High glacis pediplain in Fig. 11b. Red arrows represent drainage rearrangements
803 in between the High glacis stage (d) and the present-day landscape (a). (e) Slope map of
804 the current topography derived from (a). (f) Slope map of the High glacis pediplain
805 derived from (b). (g) Comparative slope distribution for the High glacis and modern
806 topographies. (h) River network automatically extracted from the present-day
807 topography in Fig. 11a.

808

809 **Figure 12:** Geomorphological map of the High glacis pediplain before its abandonment
810 in the Earliest Late-Miocene (ca. 11 Ma). The thick black dashed line represents the
811 boundary of the Mouhoun catchment.

812

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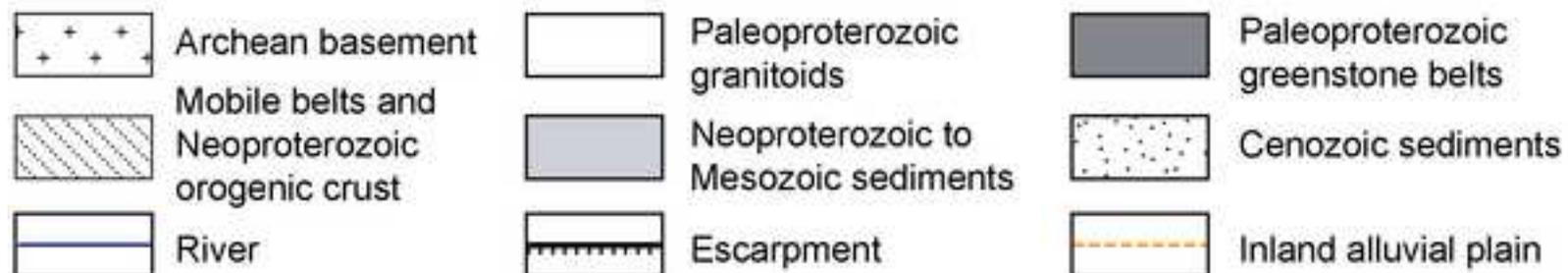
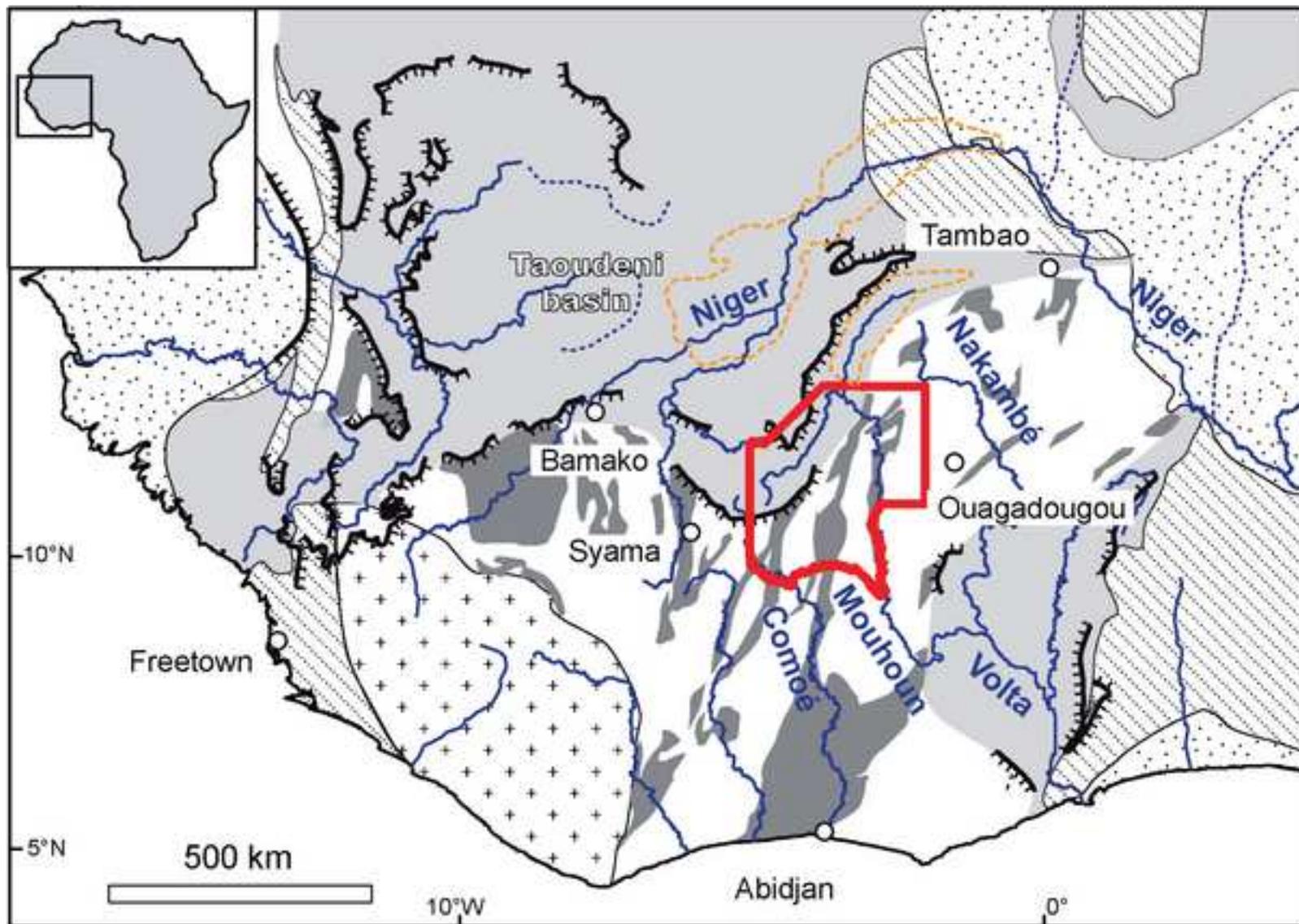


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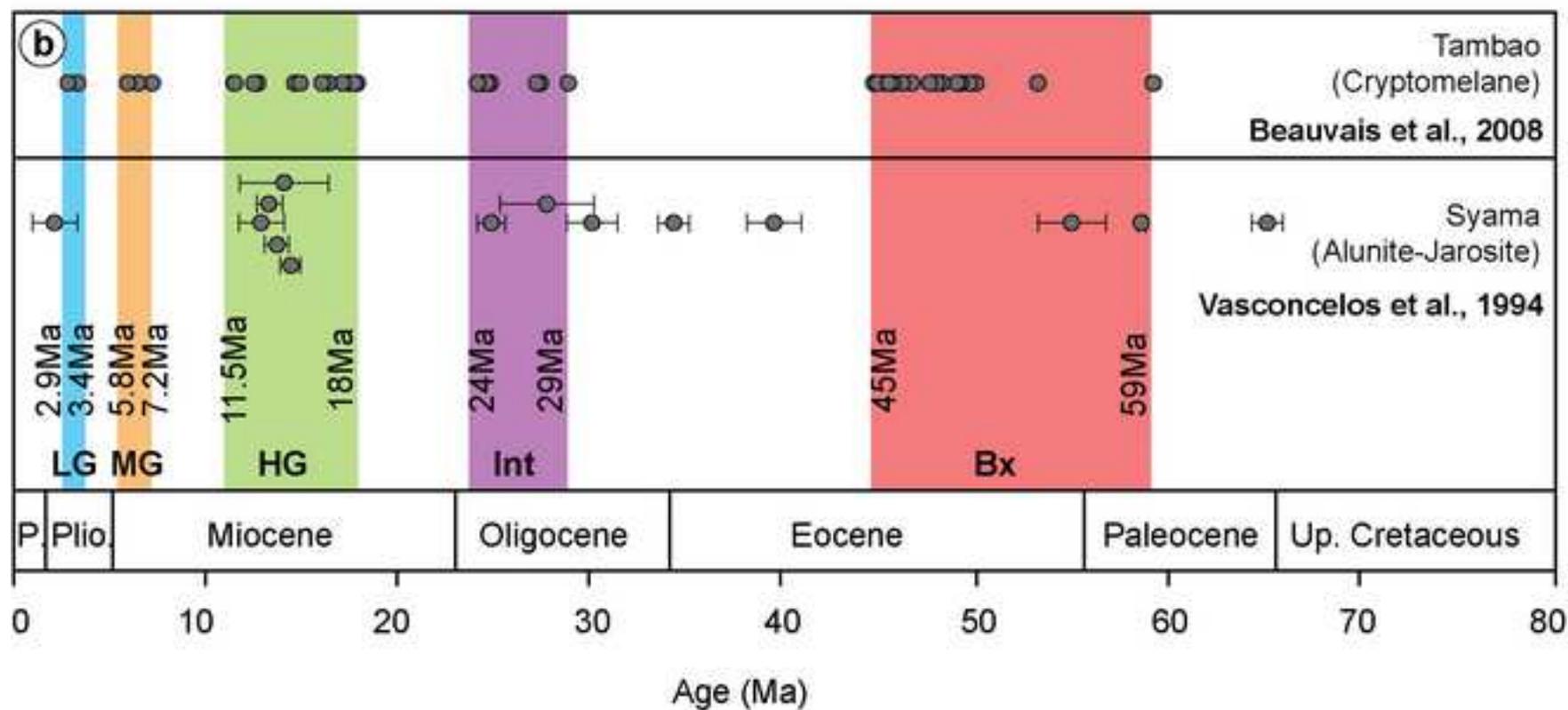
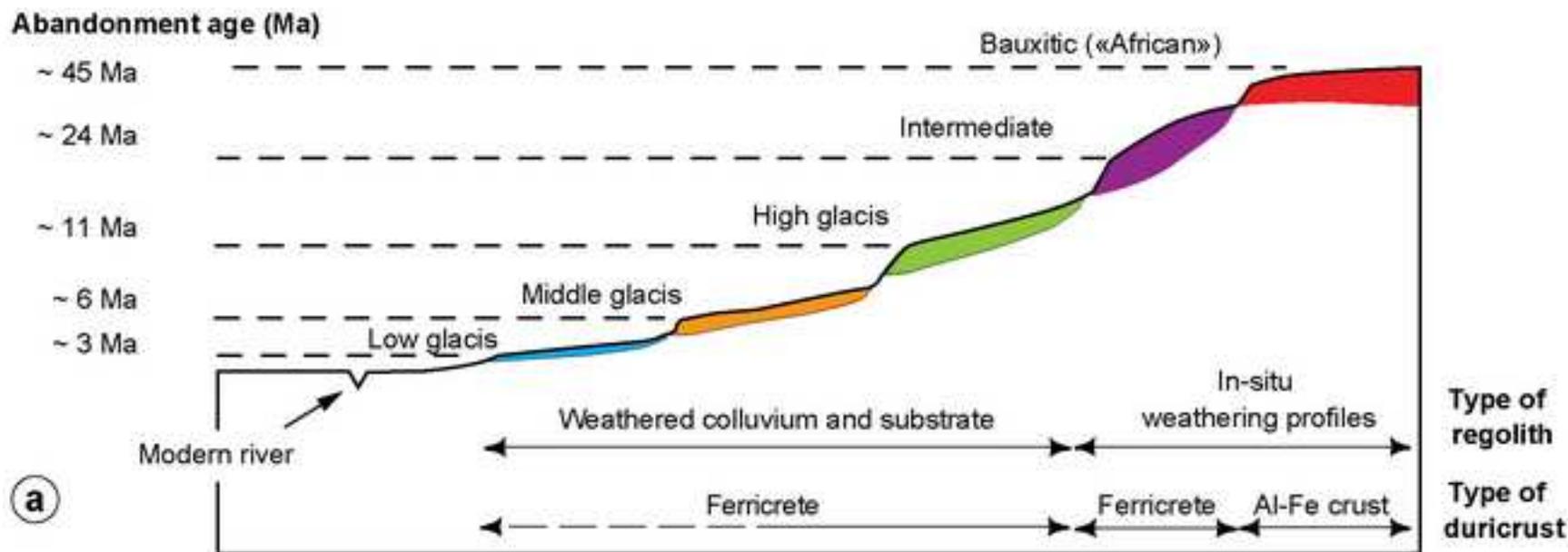


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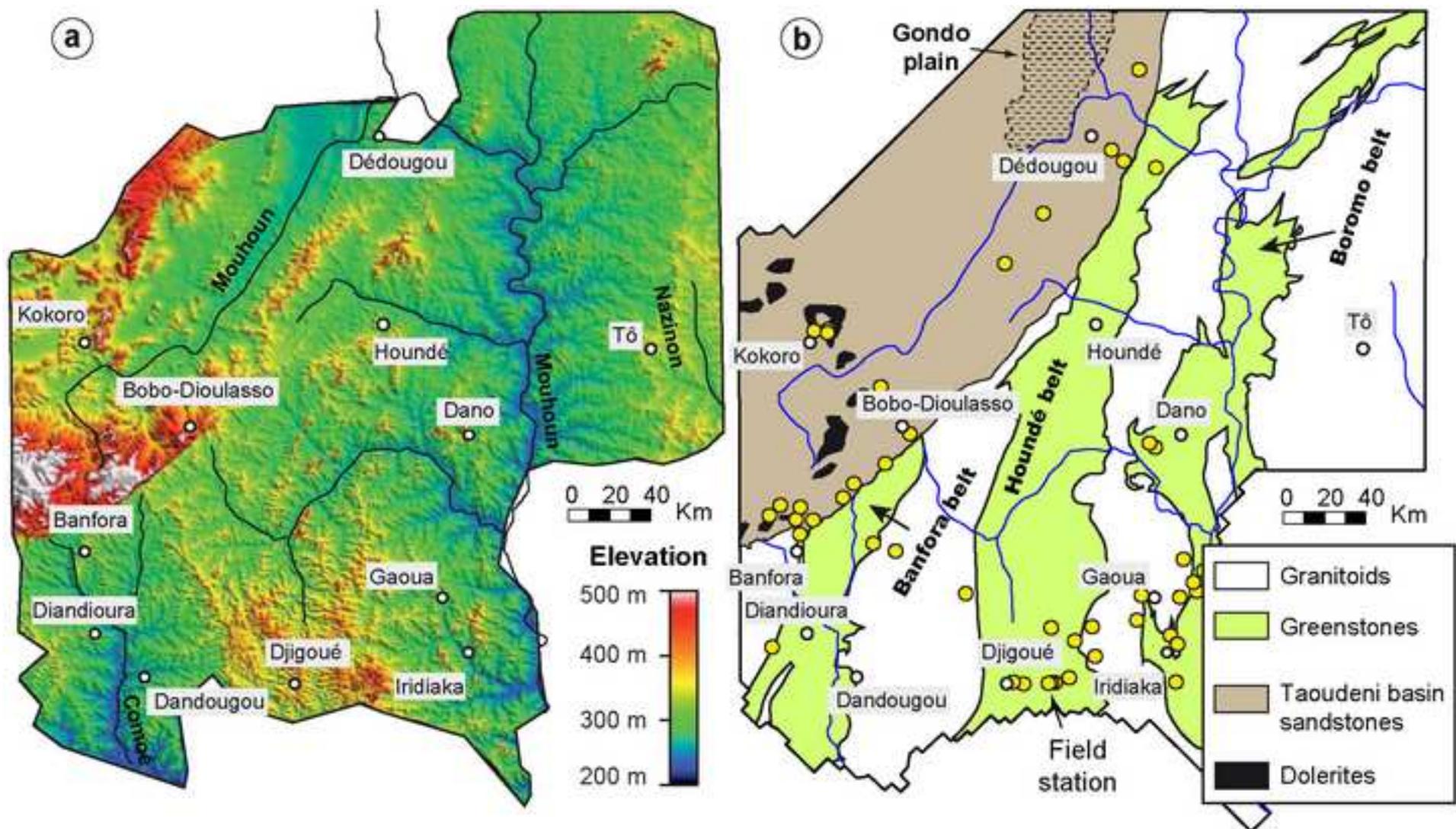


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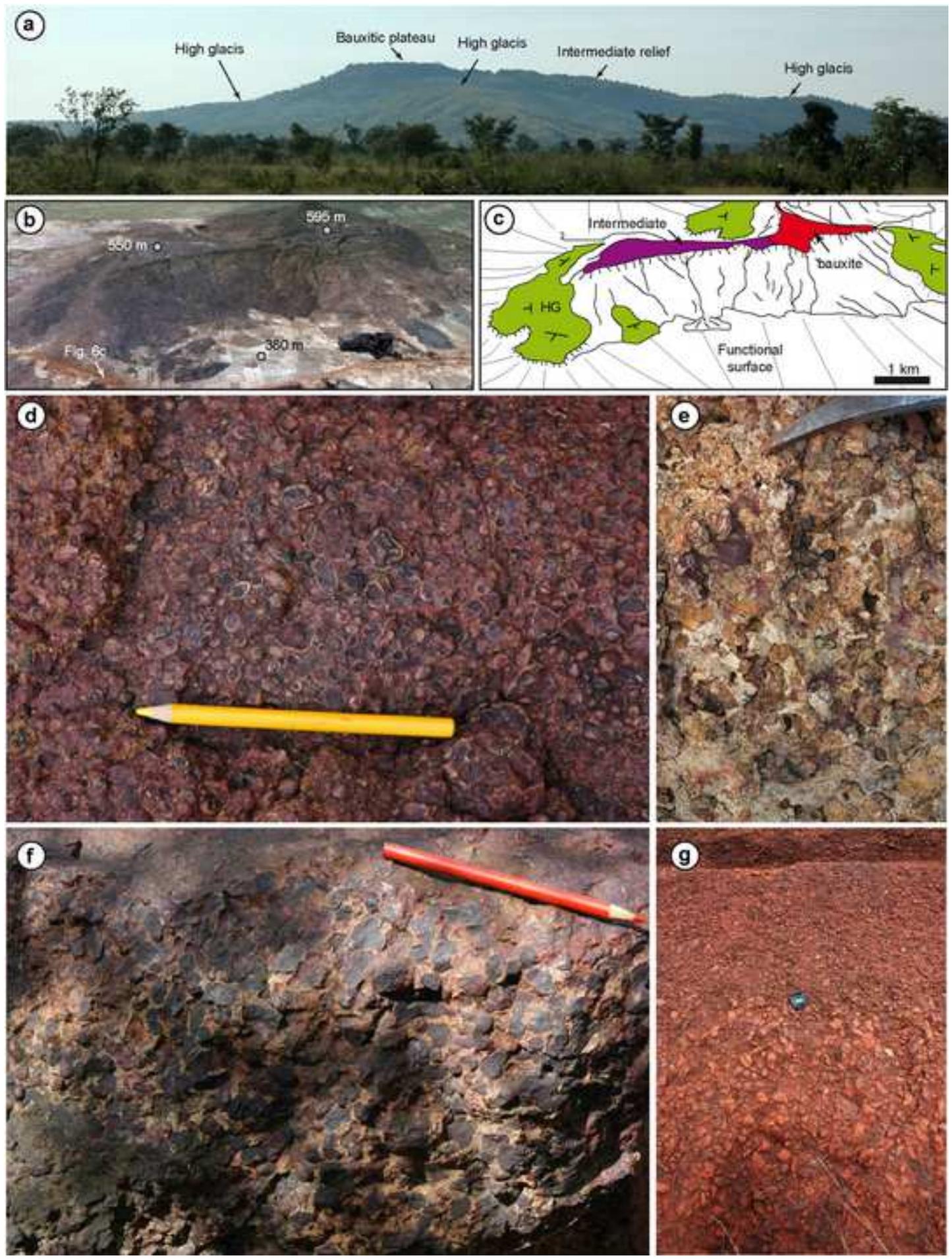


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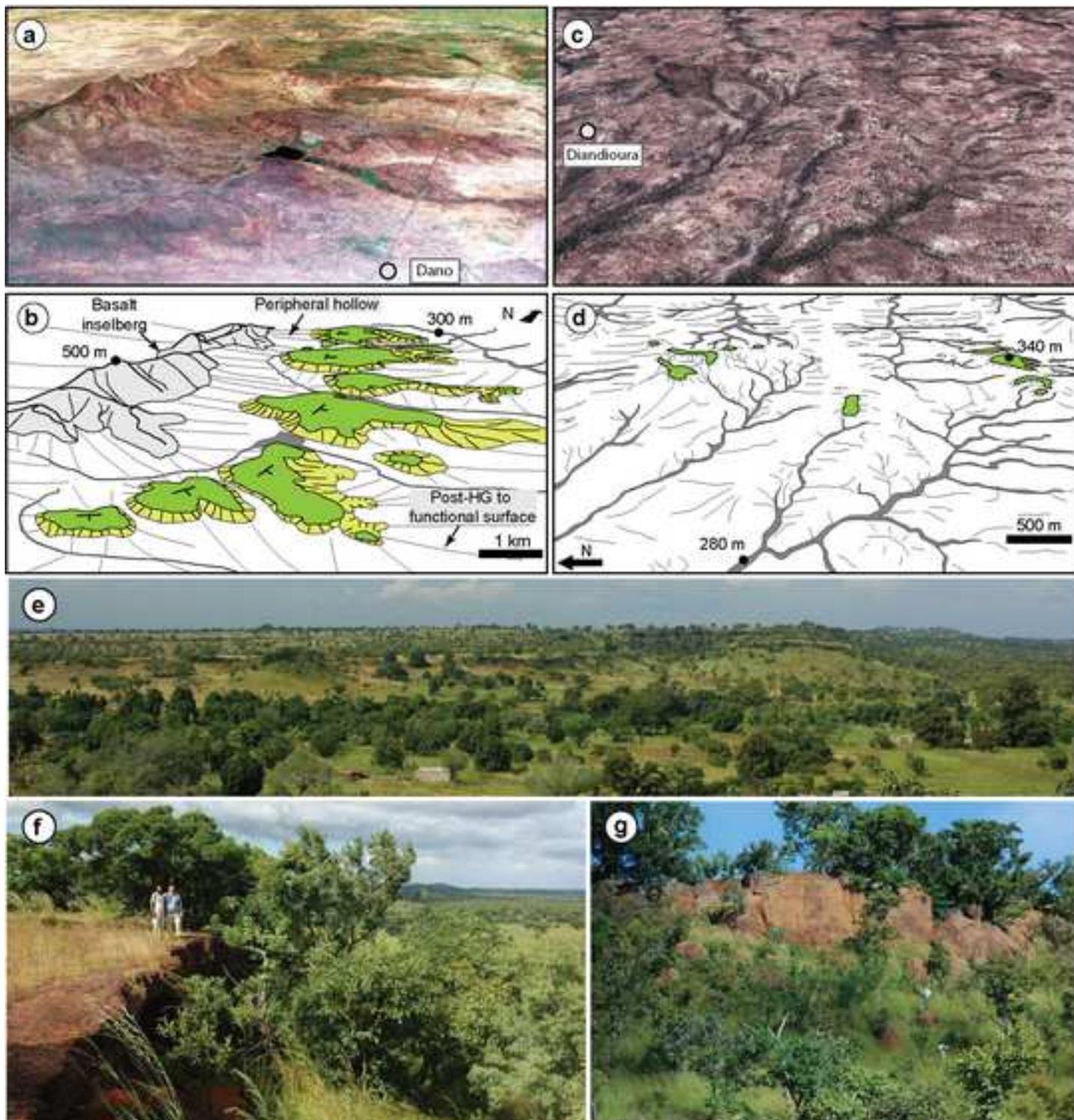


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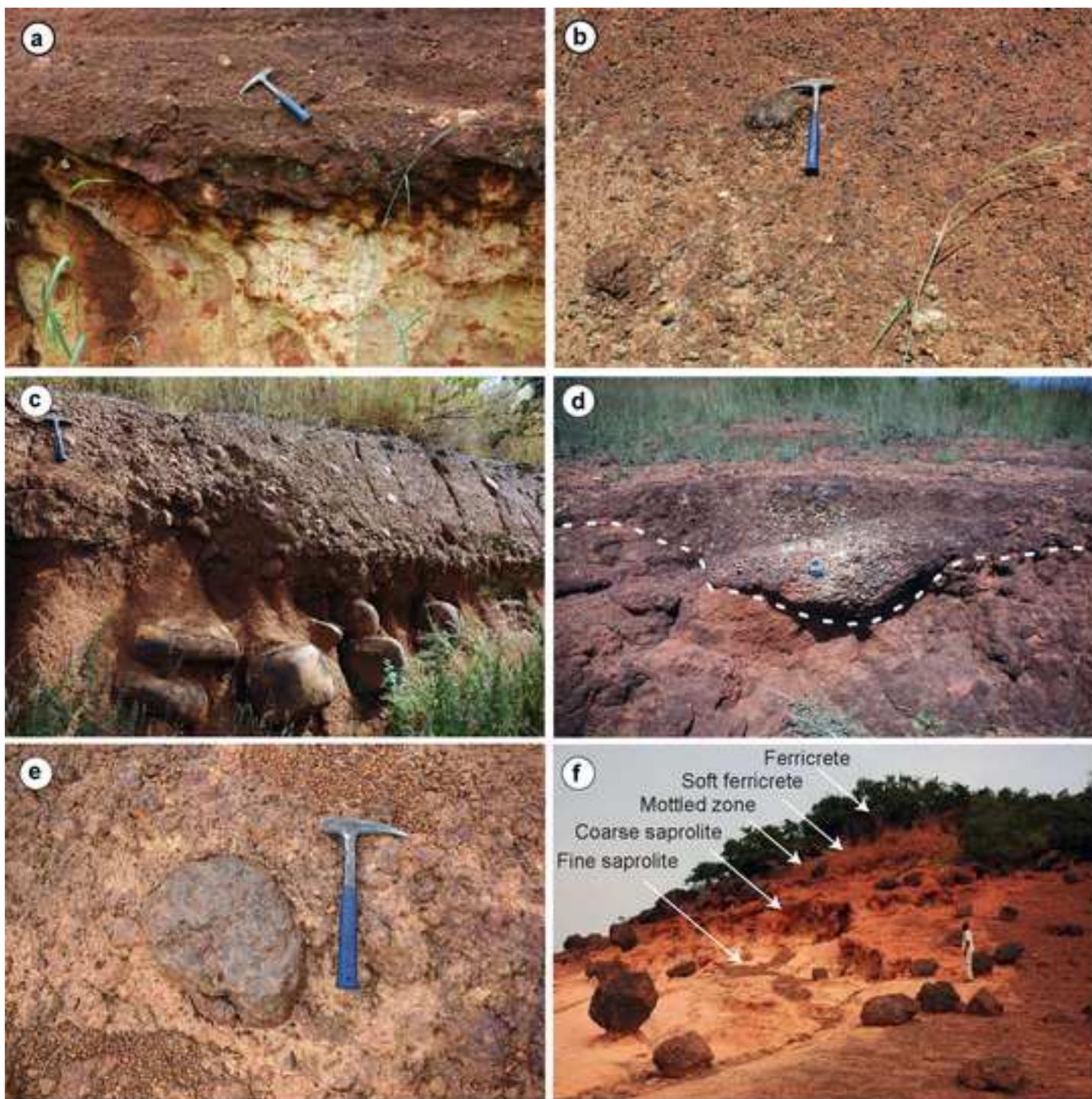


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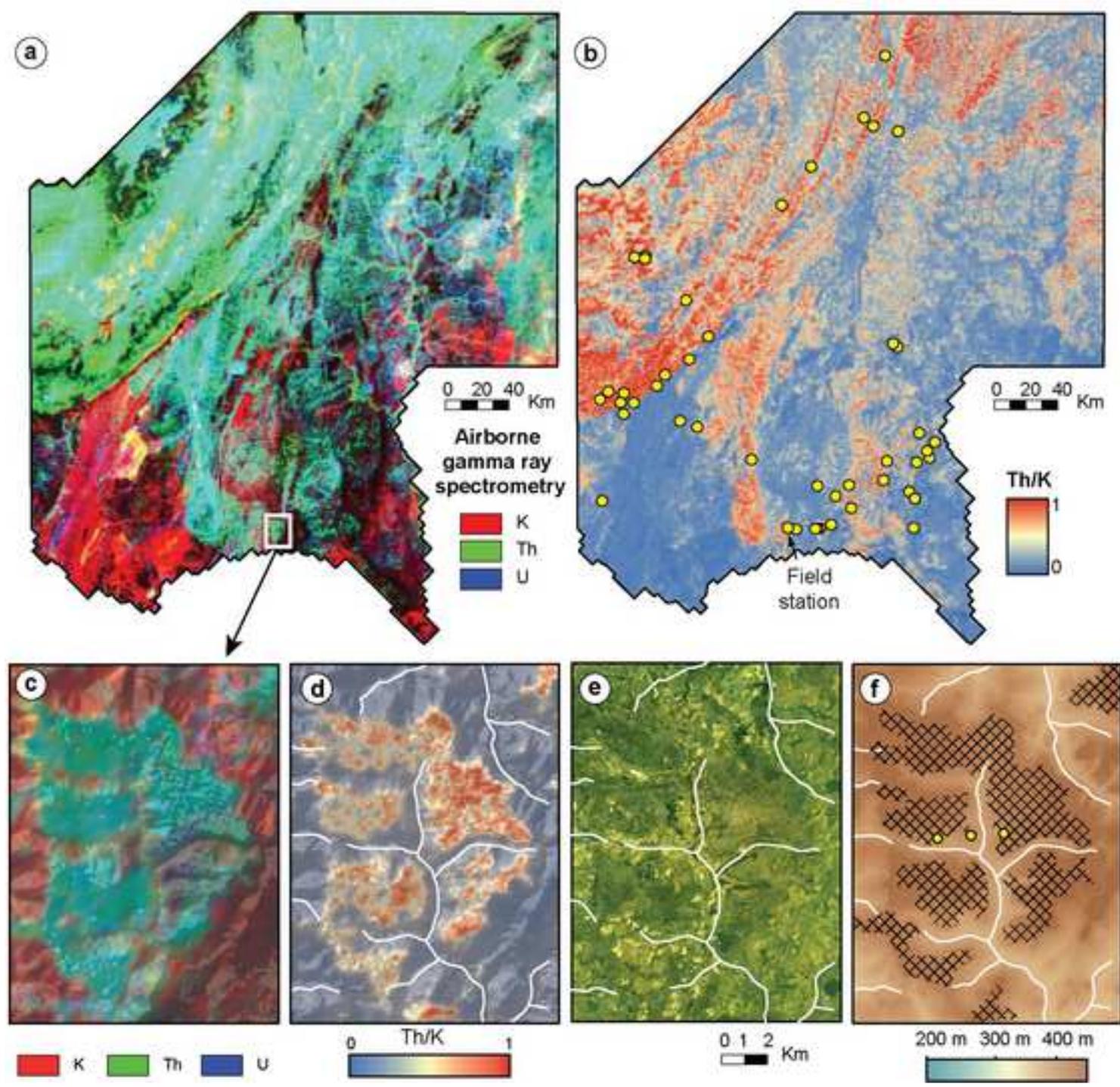


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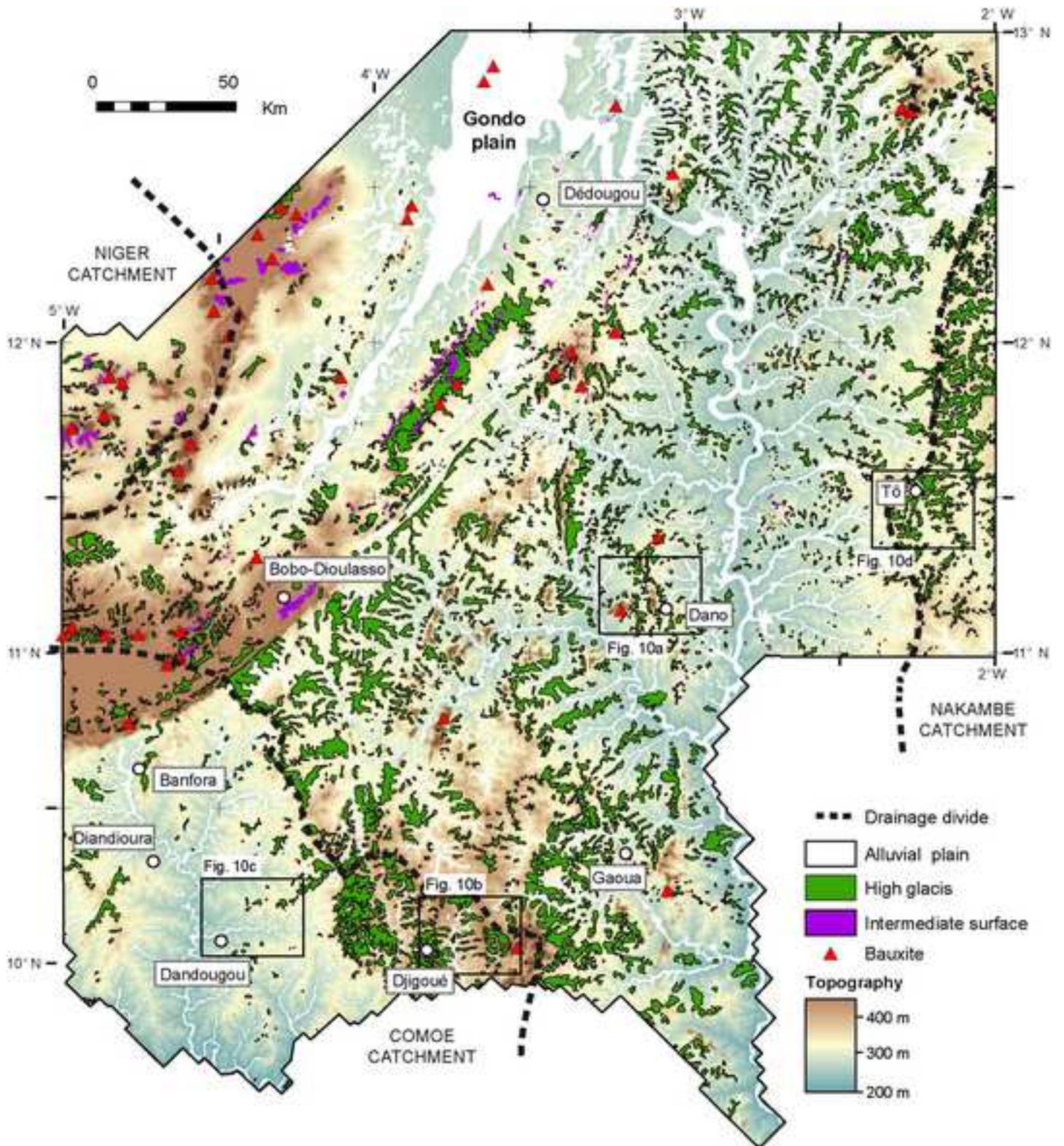


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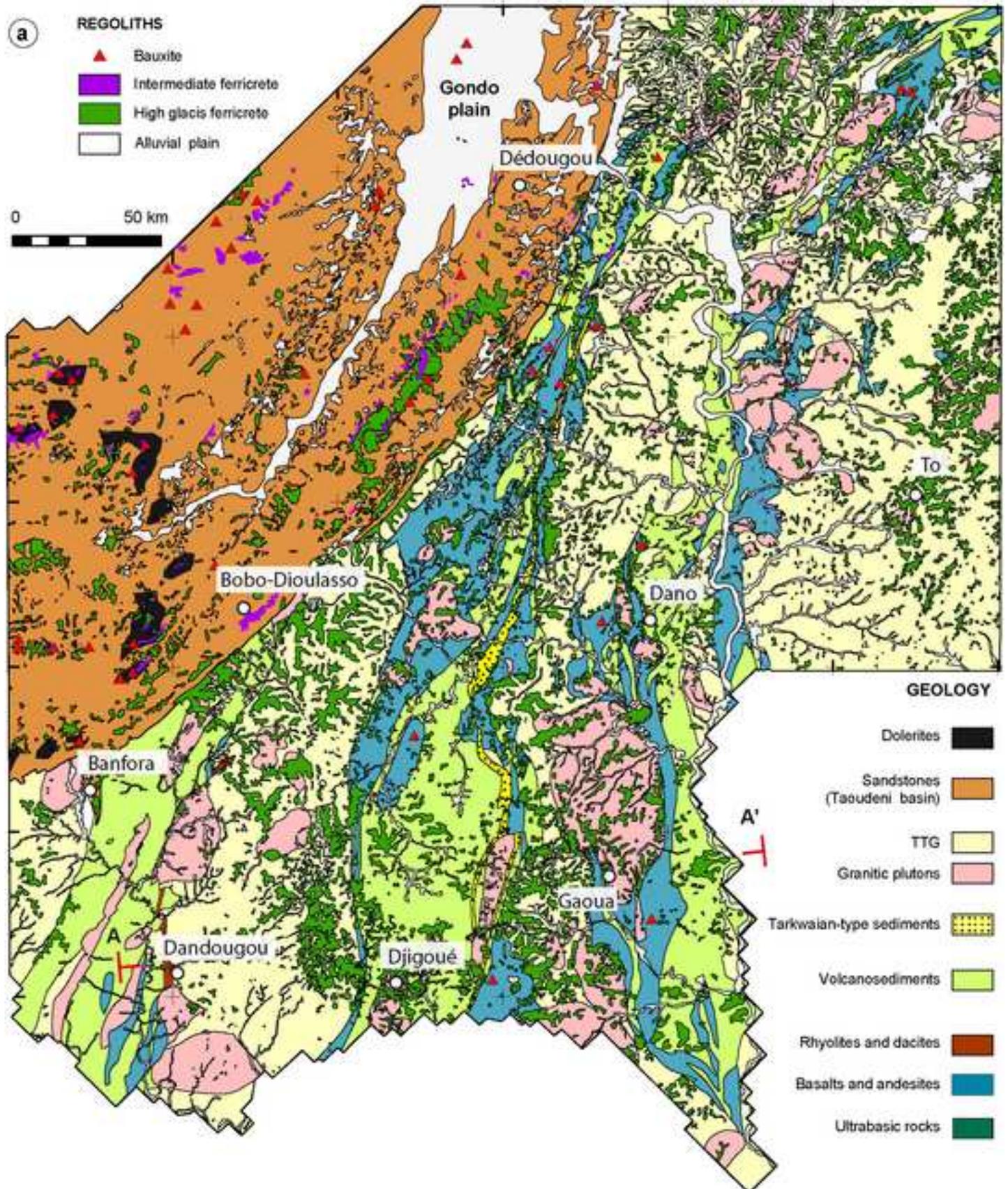


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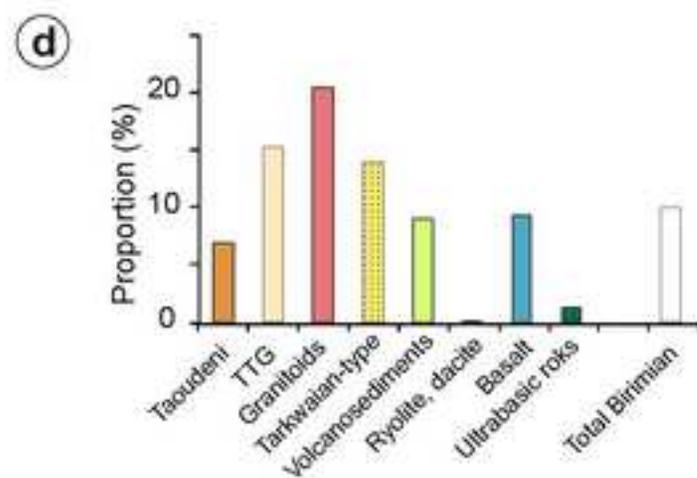
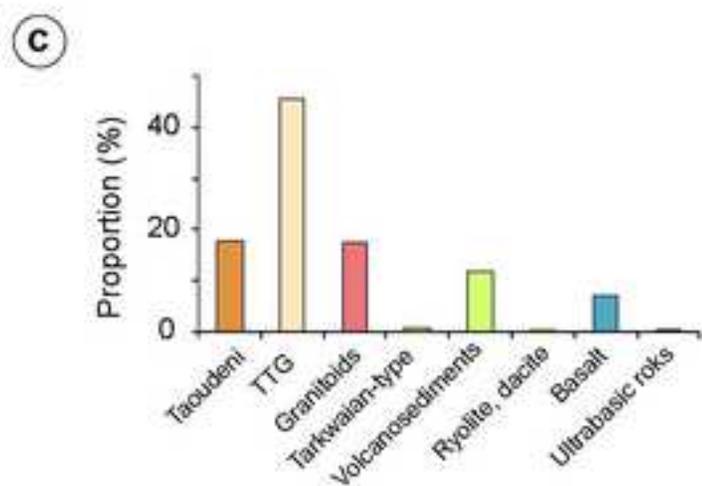
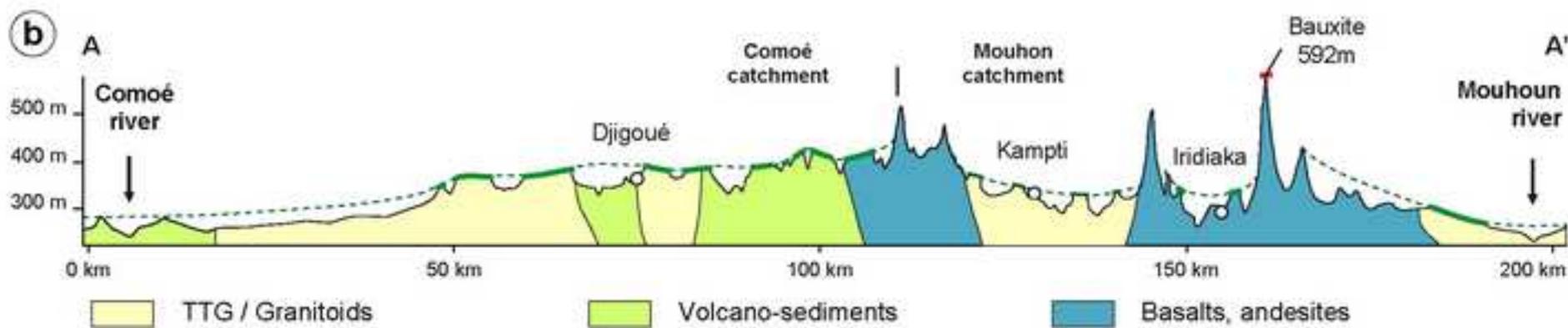


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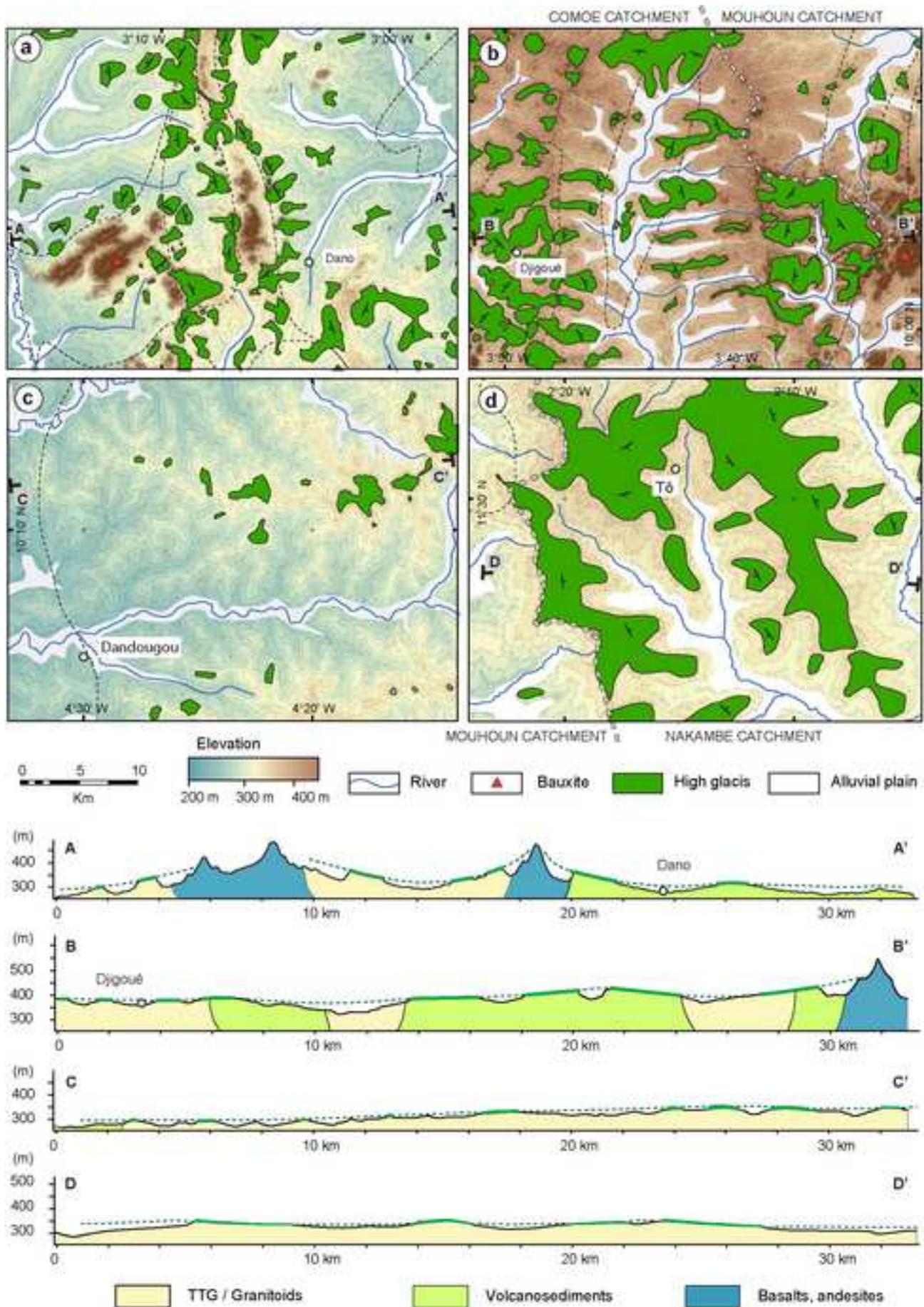


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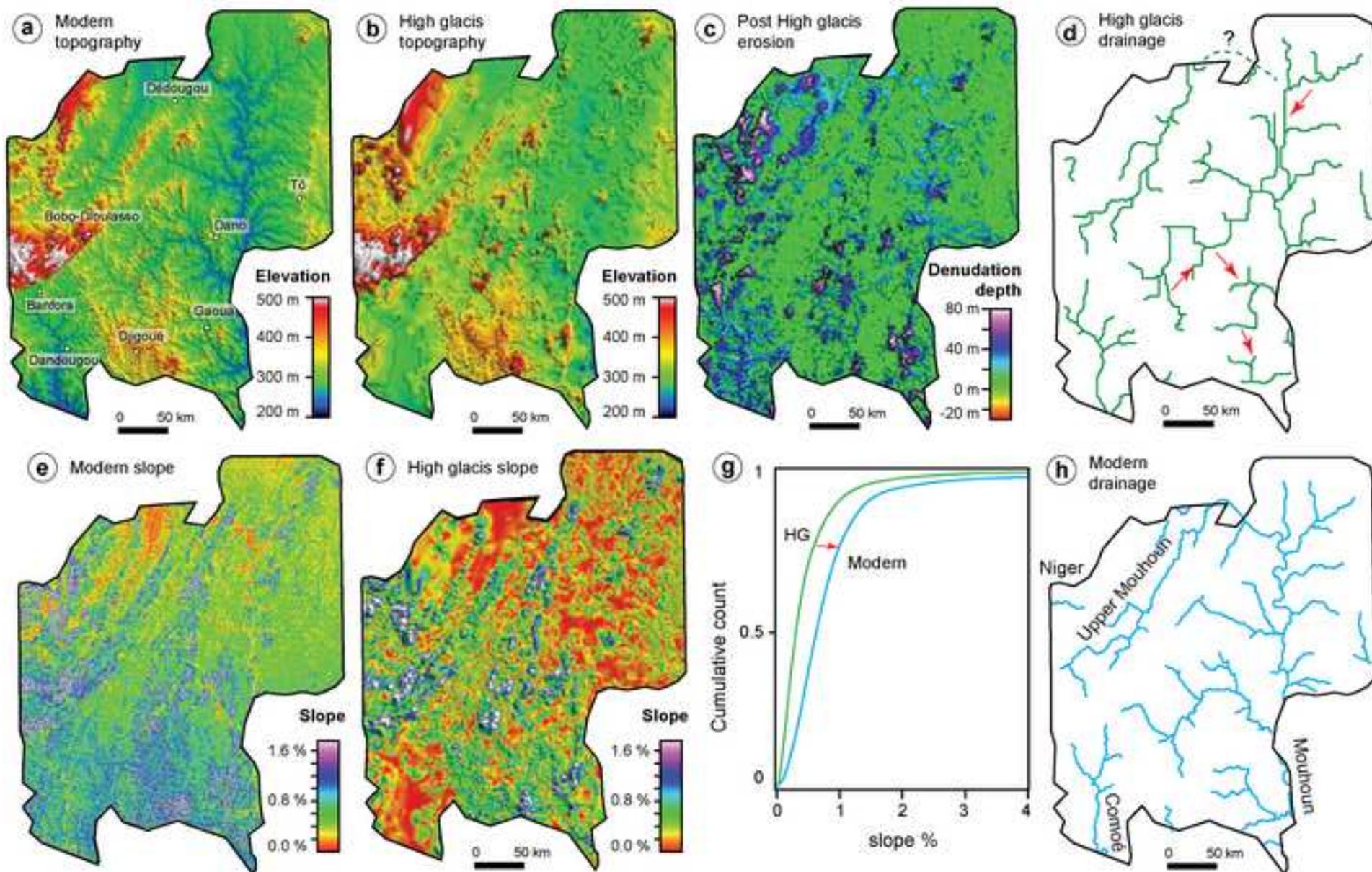


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